Rock strength along a fluvial transect of the Colorado Plateau - quantifying a fundamental control on geomorphology

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Rock strength along a fluvial transect of the Colorado Plateau – quantifying a fundamental control on geomorphology

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Abstract

Bedrock strength is a key parameter that influences slope stability, landscape erosion, and fluvial incision. Yet, it is often ignored or indirectly constrained in studies of landscape evolution, as with the K erodibility parameter in stream-power models. Empirical datasets of rock strength suited to address geomorphic questions are rare, in part because of the difficulty in measuring those rocks at Earth’s surface that are heterolithic, weak, or poorly exposed. Here we present a large dataset of measured bedrock strength organized by rock units exposed along the length of the trunk Green–Colorado River through the iconic Colorado Plateau of the western U.S. Measurements include field compressive tests, fracture spacing, and Selby Rock Mass Strength at 168 localities, as well as 672 individual tensile-strength tests in the laboratory. Tensile strength results are compared to geomorphic metrics of unit stream power, river gradient, and channel and valley-bottom width through the arid Colorado Plateau, where the influence of bedrock is intuitive but unquantified.

Our dataset reveals logical trends between tensile and compressive strength as well as between strength, rock type and age. In bedrock reaches of the fluvial transect, there is a positive rank-correlation and a strong power-law correlation between reach-averaged rock strength and unit stream power, as well as a linear relation between tensile strength and river gradient. Expected relations between fracture spacing and topography are masked partly by the massive yet weak sandstones in the dataset. To constrain values for weak rock types such as shale, we utilize the inverse power-law scaling between tensile strength and valley-bottom width to estimate their “effective” tensile strength. Results suggest that tensile strength varies to at least an order-of-magnitude smaller values than evident with directly testable rocks in this landscape, and values for erodibility (K) in numerical simulations may be informed by this dataset. In terms of landscape evolution, these results support the finding that equilibrium adjustment to bedrock strength, not differential uplift or transient incision, is the first-order control on large-scale fluvial geomorphology in the Colorado Plateau. This has broad implications for the interpretation of topography in terms of tectonic drivers.

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1. Introduction

Landscape evolution occurs through the interplay of tectonics, climate, and erosional processes acting upon the geologic substrate to shape terrain over time. There has long been agreement that the geological substrate (bedrock strength) is a first-order control in landscapes (Playfair, 1802; Gilbert, 1877), but this idea is generally couched in vague terms because of a lack of data relating rock strength to geomorphology. For example, Hack’s (1975) classic conceptual model of dynamic equilibrium in landscapes over long timespans relates the balance between bedrock resisting forces and erosional driving forces. He recognized that areas of resistant rock are higher in steepness and relief than areas of “soft” rock, even assuming that erosion rates may be uniform across the entire landscape. Bedrock strength is most explicitly studied in the context of slope stability thresholds, and landscape-evolution research has focused on the role of that in setting limits to topographic relief and form (Selby, 1980; Schmidt and Montgomery, 1995). Likewise, where rivers are in contact with varying bedrock, equilibrium theory predicts a correlation between channel steepness or stream power and bedrock strength, with a river’s gradient and width adjusted to provide the driving forces...
necessary for incision and maintenance of baselevel (Mackin, 1948; Whipple and Tucker, 1999). In the case of the Colorado Plateau, John Wesley Powell (1895) recognized very early a connection between old, resistant rock units and steep, treacherous gorges along the Colorado River:

“A river may be hundreds of miles in length. As it flows along, it passes through rocks of varying degrees of hardness... In this manner the river is divided into lengths, or reaches. Along its course where the rocks are hard, the stream is narrow and swift, with rapids and falls; where the rocks are soft, it is wide and quiet.”

Despite these basic expectations, most modern research in tectonic geomorphology and landscape evolution has effectively ignored or downplayed bedrock strength as an important control. It is not clear which data are specifically applicable to surface processes, such as compressive or tensile strength, fracturing, or some combination of these. In measuring rock strength in geomorphic context there is also an issue of sampling bias. We are limited to rock units that outcrop and are not too fractured, thinly bedded, or weakly indurated to withstand field and laboratory measurements. Finally, how to integrate such data into landscape evolution models is poorly developed (Pazzaglia, 2003; Tucker and Hancock, 2010).

In research on the erosion of bedrock streams, Stock and Montgomery (1999) explored the influence of bedrock on fluvial incision by deriving the erodibility factor ($K$) of the stream power law while modeling river-profile evolution. Subsequent numerical, laboratory, and field studies have built upon this, recognizing, for example, that channel and valley-bottom width decreases in reaches of resistant bedrock, sometimes without changes in gradient (Montgomery and Gran, 2001; Montgomery, 2004; Limaye and Lamb, 2014). Yet, varying bedrock type along a river’s length also creates knickpoints (Miller, 1991; Goldrick and Bishop, 1995), with the baselevel signal diffusing quickly through channels in weak rock units and remaining focused on resistant rocks as it migrates upstream (Gardner, 1983; Berlin and Anderson, 2007; Cook et al., 2009). Complicating the relation between rock strength and fluvial incision is the effect of the grain size and amount of sediment load, either used as tools for incision or covering the bed and preventing it (e.g. Howard, 1998; Whipple and Tucker, 2002; Sklar and Dietrich, 2006; Johnson et al., 2009). Yanites and Tucker (2010) find through numerical modeling that both the width and gradient of a bedrock river should scale as a changing function of sediment load and bed cover. These findings indicate that unraveling the complex controls on the incision of bedrock streams requires not only actual measurements of rock properties, but also knowledge of sediment transport and supply.

Our case study of the erosion of the Colorado Plateau involves an unresolved mix of different sources of baselevel fall/uplift and transient incision. Although there is ongoing debate about the history of various segments of Grand Canyon (Wernicke, 2011; Karlstrom et al., 2014), the Colorado River as we know it was integrated across the southwestern edge of the Colorado Plateau at ~6 Ma, causing a ~1500 m baselevel fall (Fig. 1; Pederson et al., 2002). This baselevel signal has propagated upstream through this transient landscape, setting into motion isostatic feedbacks (Cook et al., 2009; Pederson et al., 2013). In addition, there are hypothesized sources of differential uplift, or broad tilting, along the flanks of the Colorado Plateau from mantle effects (e.g. Mouha et al., 2009; Levander et al., 2011).

Our approach is to test the hypothesis that the Colorado River’s profile is in dynamic equilibrium with reach-scale bed resistance imparted directly by bedrock or indirectly by the bedload it supplies (Pederson and Tressler, 2012). If so, then unit stream power, width and gradient of those reaches should correlate with rock strength in our dataset. If these correlations do not exist or have exceptions in particular reaches, then differential uplift or disequilibrium, transient conditions may dominate.

To more broadly address the influence of bedrock strength in landscape evolution, we present a comprehensive dataset of multiple measures of rock strength along the trunk Green–Colorado River drainage through the Colorado Plateau (Fig. 1). We focus on tensile strength and on bedrock reaches rather than alluvial reaches, and then explore basic relations between those data and reach-scale stream power and fluvial-topographic metrics. We utilize a power-law correspondence of valley width to bedrock tensile strength to estimate “effective” tensile strength values for the weak bedrock types that cannot be directly tested. We offer this dataset as an empirical asset for colleagues to further utilize.

2. Background

2.1. Bedrock-strength measures

Multiple approaches exist to quantify rock strength in geomorphology. For hillslopes, strength analysis stems from Mohr–Coulomb failure, relating cohesion and frictional resisting forces to gravitational driving forces, pore pressure, and fractures, and threshold hillslopes have been related to the limits of topographic relief (Schmidt and Montgomery, 1995). The Selby Rock Mass Strength (RMS) semi-quantitative field classification is also used to capture hillslope stability, including compressive-strength measurements along with other observations of fracturing and weathering (Selby, 1980). Researchers have modified the Selby RMS according to their applications, especially attempting to capture the...
strength of weak rocks (Moon et al., 2001; Brook and Hutchinson, 2008). The compressive strength of rocks, including for Selby RMS, is typically estimated using a Schmidt hammer in the field. Measurements must be far enough from joints and bedding planes to get a correct reading of inherent elasticity (Young’s Modulus), and consequently shale and other thinly bedded rocks cannot be tested. A notably different approach focused on fracturing is using seismic (P-wave) velocities to quantify depth-dependent variations and mechanical properties (Clark and Burbank, 2011).

For establishing erodibility with respect to fluvial systems, researchers have used Schmidt hammer measurements alone, or as incorporated into Selby RMS, somewhat ignoring fracturing and any thinly bedded rocks (e.g., Wohi and Merritt, 2001; Duvall et al., 2004). Yet, incision into bedrock by streams occurs through breakage in tension, not compression, and it is strongly influenced by bedrock fracturing (Whipple et al., 2000). The physical experiments of Sklar and Dietrich (2001) have shown that tensile rock strength is correlated with erosion rate, by abrasion specifically. Our data for this study includes Selby RMS and Schmidt-hammer compressive strength, but we focus our analysis on tensile-strength measurements for these reasons.

Tensile strength for geomorphic studies has been determined by the Brazilian splitting test in the lab (Yutkuri et al., 1974). The standard Brazilian test utilizes 25.4 mm (1 inch) thick and 50.8 mm (2 inch) diameter discs, and a uniaxial stress is applied until a primary fracture forms parallel to the loading vectors. The load (p in N) required to induce failure is used to calculate the tensile strength ($\sigma_t$, in MPa) of the rock using:

$$\sigma_t = \frac{2p}{\pi LD},$$

where L (mm) and D (mm) are the length and diameter of the discs being tested. Similar to Schmidt hammer measurements, the Brazilian splitting test is limited to rocks that are coherent enough to be transported, cored, and made into discs. From theory, given uniform material, it is suggested that compressive strength should simply be twice the tensile strength (Jaeger and Hoskins, 1966a). However, this assumption is in contrast with experimental results from a variety of natural rock types as shown by our data (SM Fig. 1) and those in rock-mechanics datasets (e.g. Jaeger and Hoskins, 1966b; Lockner, 1995). Furthermore, significant scatter is expected in empirical data because of natural rock variability and contrasting measurement methods.

2.2. Bedrock erodibility in fluvial geomorphology

The stream power erosion law is often used to examine the work done by bedrock streams, especially for modeling landscape evolution. It describes the rate of incision ($E$) as a function of the scalar $K$, and contributing drainage area ($A$) used as a proxy for discharge and channel gradient ($S$), raised to the power of $m$ and $n$, respectively:

$$E = KA^mS^n.$$  

The constants $m$ and $n$ tend to be set as prescribed from mathematical derivation. The dimensional coefficient $K$ represents overall erodibility and the effects of bedrock strength, but also incorporates climate and runoff efficiency, channel-width scaling, and sediment load (Howard and Kerby, 1983; Whipple and Tucker, 1999).

Bedrock resistance is central to setting the $K$ erodibility, but understanding which rock properties and what processes govern the rate of erosion remains a major research question. Numerical modeling studies typically estimate $K$ from topography or adjust values to yield model results that resemble nature rather than using direct measurements (Howard et al., 1994; Stock and Montgomery, 1999; Whipple and Tucker, 1999; Attal et al., 2008). In an especially pertinent example simulating the erosion of Grand Canyon, Pelletier (2010) divided the canyon’s stratigraphy into two $K$-value groupings such that model results best matched the profiles of tributary drainages.

Fluvial incision into bedrock occurs by abrasion, plucking, cavi
tation and solution, with plucking dominant on well-jointed, bed-
ded or fractured rocks and abrasion dominant in massive rocks (Whipple et al., 2000). All these processes (except solution) in-
volve rocks breaking in tension, suggesting that tensile strength 
along with fracturing should be of primary importance. Sklar and 
Dietrich (2001) established that erosion of rock by abrasion scales with the square of tensile strength. Furthermore, sediment supply in providing tools and in covering the bed and controlling exposure of rock is an important factor, and modeling indicates that igno-
ring these tools and cover effects results in inaccurate predictions of incision rate and channel gradient (Sklar and Dietrich, 2001; Gasparini et al., 2007).

In the Colorado Plateau, there is a variety of massive and bed-
ded or fractured rocks, including metamorphic, sedimentary, and 
intrusive and extrusive igneous units (Figs. 1 and 2). The domi-
nant fluvial incision process must change from plucking in highly 
fractured rocks to abrasion in massive rocks from reach to reach. 
The Colorado River is also a mixed bedrock–alluvial river (Howard, 1998), and bedrock gorges alternate with alluvial valleys where the river has not been in contact with bedrock during the Holocene. Though the river currently does not act directly upon bedrock in such reaches, including some in Grand Canyon, it has obviously done so over longer geologic history, and its coarser bedload reflects local lithology (cf. Hanks and Webb, 2006). Yet, the natural sediment load of this river system is high, and bed-cover effects may provide a further complication in this system.

3. Research design

Our data are collected along the length of the trunk drainage 
starting in Wyoming where the Green River crosses the struc-
tural basin of the Paleogene Green River Formation (Figs. 1 and 2). Flowing south, the river enters the Uinta Mountain knickzone first through Red Canyon’s Proterozoic Uinta Mountain Group sand-
stone, then the Neogene Browns Park heterogeneous and tuf-
faceous basin fill, and then the same Uinta Mountain Group and also Paleozoic sedimentary rocks in the canyons of Dinosaur National Monument (SM Fig. 2, Fig. 2). The Green River then enters the shale-rich upper Mesozoic and Cenozoic strata of the Uinta Basin and the Desolation Canyon knickzone. Mesozoic sedimentary rocks dominate the river corridor through the central Colorado Plateau, but just below the confluence of the Green and Colorado Rivers, it plunges through a window of Paleozoic rocks in the short, steep Cataract knickzone. The Grand Canyon knickzone, at the down-
stream end of the Colorado Plateau, is carved in Paleozoic and 
Proterozoic strata, eventually encountering Proterozoic basement 
rocks before debouching into the Basin and Range (Fig. 2).

3.1. Bedrock strength

We divided this river pathway into forty-nine distinct bedrock-
egology reaches, updating those previously defined by Pederson 
and Tressler (2012) such that our reach lengths, locations and channel metrics deviate slightly from theirs. Thirty-two of the

fourty-nine reaches are considered bedrock reaches where the river is at least somewhat confined to canyons and occasionally in con-
tact with rock (Fig. 2). The seventeen other reaches are considered alluvial, where the channel occupies its own floodplain through open valleys. These alluvial reaches coincide with bedrock units that contain greater than 50% shale or mudstone. We focus on
bedrock reaches for much of our analysis because our goal is to explore how bedrock may influence the form of the river and its canyons. Yet, we presume that alluvial reaches also have overall valley-bottom widths (distinct from channel widths) that reflect geologic controls, inasmuch as the shifting channel has laterally carved that valley into bedrock over geologic time.

Selby RMS classification, including fracture spacing and Schmidt hammer measurements, was completed for 52 named geologic formations at 168 outcrop localities (Table 1, see SM Tables 1A–1C for full data). Forty-one formations were sampled for laboratory tensile-strength testing from 64 of these outcrops. Sampling locations were selected for accessibility by road or river and to be representative of the formations encountered by the river along the main-stem corridor. Of the 49 reaches in this study, over half include data for more than one bedrock unit, and reach-average values were used for most comparisons. Only reaches 18 (Mancos shale in Gunnison Valley) and 29 (Paradox Formation evaporites in upper Cataract Canyon) have no rock-strength data of any kind. Likewise, we take into account that the data for alluvial reaches 8, 10, 11, 19 and 22 are from the resistant beds within those shaley units and are consequently unrepresentative.

Compressive strength values were measured using a Schmidt hammer in the field, with 50 measurements taken for each locality. The raw rebound values are utilized in the overall Selby RMS classification, but Schmidt hammer measurements were also converted into a compressive strength value in MPa using the formula empirically derived for the instrument:

$$C = 2.12 \cdot R^{1.06}.$$  \hspace{1cm} (3)

Samples for tensile-strength were prepared and measured at Utah State University following procedures for the Brazilian splitting test as described above. An average of 10 discs for each rock formation were load-tested to their yield strength.

3.2. Topographic metrics

Along the Colorado and Green River trunk drainage, at 0.5 km nodes, we computed or measured gradient, channel width, valley width, and discharge (SM Table 2). Where digital topography is obscured by the Flaming Gorge and Lake Powell reservoirs, terrain was digitized from the contour lines of older maps that display pre-dam topography (Pederson and Tressler, 2012). To avoid the errors and artifacts of DEM-derived channel profiles, elevations along the drainage were digitized from pre-dam USGS survey data and used to calculate gradient (Pederson and Tressler, 2012). Channel width was measured at each node directly from the channel shape as recorded on 1:24,000-scale topographic maps. For valley-bottom width, the Barr-NCED Flood Inundation Tool (http://www.nced.umn.edu/content/stream-restoration-toolbox) was used to “flood” the valley to an elevation of 10 m and 50 m above present river level. Valley width at each node was measured directly from the flood polygons produced.

A plot of the reach-averaged channel widths versus the 50-m-height valley widths highlights the contrast between bedrock and alluvial reaches (Fig. 3). Although they share a similar range in channel widths, bedrock reaches are less than 1000 m in valley width, whereas alluvial reaches range beyond 4000 m. There is a positive correlation between channel and valley width in bedrock reaches, but no such scaling exists in alluvial reaches where channel width is instead a function of alluvial hydraulic geometry. With our exploration of possible bedrock controls, we focus on comparing the valley width of bedrock reaches to varying bedrock strength, as lateral planation of the river has carved valley bottoms over geologic time.

To estimate an appropriate discharge for calculating stream power, we determined the 2.5-yr flood recurrence interval based upon pre-dam historic records at each gauging station along the rivers. Effective discharge values were plotted against contributing area from a flow accumulation raster and a least-squares regression was used to model effective discharge at any point along the length of the drainage. These discharges were used with the measured channel width and gradient to calculate unit stream power, a river’s rate of energy expenditure per unit area of bed (Table 1). Although following a similar pattern, our unit stream power values differ from those of Pederson and Tressler (2012), who calculated it using valley (not channel) width.

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**Fig. 2.** A) Longitudinal profile of the Green–Colorado River through the Colorado Plateau with bedrock exposed at river level and in canyon walls depicted above grade, and river miles relative to Lee’s Ferry. The 49 study reaches are partly labeled along the bottom axis, and major canyon knickzones, separating alluvial valleys, and physiographic features are labeled, after Pederson and Tressler (2012). Histograms are of reach-average B) unit stream power; C) tensile (above) and compressive strength (below) in MPa, with reaches 27 and 28 not included in statistical analysis (clear) and reaches 8, 10, 11, 13, 15, 24 and 25 transparent because they illustrate only the strength of resistant beds within mostly shale formations; and D) fracture spacing in meters as ranked in bins for Selby rock-mass strength.
## Table 1
Reach-scale rock-strength and hydromorphic data.

<table>
<thead>
<tr>
<th>Reach Name</th>
<th>Length (km)</th>
<th>River mile&lt;sup&gt;a&lt;/sup&gt;</th>
<th>Bedrock/Formation sampled</th>
<th>Compress. strength (MPa)</th>
<th>Tensile strength (MPa)</th>
<th>Effective tensile strength&lt;sup&gt;b&lt;/sup&gt;</th>
<th>Shale % of reach&lt;sup&gt;c&lt;/sup&gt;</th>
<th>Alluvial/bedrock</th>
<th>Channel width (m)</th>
<th>Valley width (m)&lt;sup&gt;d&lt;/sup&gt;</th>
<th>Gradient (m/m)</th>
<th>Unit stream power (W/m²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Green River Basin</td>
<td>116.8</td>
<td>−728</td>
<td>Green River Fm shl &amp; sltst</td>
<td>90</td>
<td>1.09</td>
<td>0.13</td>
<td>75%</td>
<td>alluvial</td>
<td>116</td>
<td>1847</td>
<td>0.0006</td>
<td>23</td>
</tr>
<tr>
<td><strong>Eastern Uinta Mtns</strong></td>
<td></td>
<td></td>
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<td></td>
<td></td>
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<td></td>
<td></td>
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</tr>
<tr>
<td>Red Canyon</td>
<td>68.5</td>
<td>−656</td>
<td>Uinta Mountain Group sst</td>
<td>163</td>
<td>3.96</td>
<td>n/a</td>
<td>20%</td>
<td>bedrock</td>
<td>76</td>
<td>562</td>
<td>0.0018</td>
<td>127</td>
</tr>
<tr>
<td>Topaz Park</td>
<td>55.4</td>
<td>−613</td>
<td>Br PK tuffaceous basin fill</td>
<td>69</td>
<td>0.80</td>
<td>0.08</td>
<td>80%</td>
<td>alluvial</td>
<td>96</td>
<td>2226</td>
<td>0.0007</td>
<td>38</td>
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<td>Echo Park</td>
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<td>−563</td>
<td>Weber sst</td>
<td>140</td>
<td>2.11</td>
<td>n/a</td>
<td>0%</td>
<td>bedrock</td>
<td>69</td>
<td>514</td>
<td>0.0012</td>
<td>124</td>
</tr>
<tr>
<td>upper Dinosaur</td>
<td>7.7</td>
<td>−559</td>
<td>Uinta Mountain Group sst</td>
<td>163</td>
<td>8.43</td>
<td>n/a</td>
<td>5%</td>
<td>bedrock</td>
<td>49</td>
<td>429</td>
<td>0.0028</td>
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<td>Madison sst</td>
<td>148</td>
<td>9.37</td>
<td>n/a</td>
<td>25%</td>
<td>bedrock</td>
<td>46</td>
<td>441</td>
<td>0.0016</td>
<td>213</td>
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<tr>
<td>Island Park</td>
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<td>Morrison (Stump) Fm sst</td>
<td>96</td>
<td>n/a</td>
<td>0%</td>
<td>60%</td>
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<td>139</td>
<td>1631</td>
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<td>Split Mountain Canyon</td>
<td>11</td>
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<td>148</td>
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<td>n/a</td>
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<td>bedrock</td>
<td>53</td>
<td>492</td>
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<td><strong>Uinta Basin</strong></td>
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<tr>
<td>Jensen</td>
<td>15</td>
<td>−537</td>
<td>Morrison Fm sst</td>
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<td>0.33</td>
<td>50%</td>
<td>alluvial</td>
<td>81</td>
<td>1270</td>
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<td>58</td>
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<td>Mancos shl</td>
<td>140</td>
<td>2.11</td>
<td>n/a</td>
<td>0%</td>
<td>alluvial</td>
<td>157</td>
<td>3830</td>
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<td>14</td>
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<td>mid Uinta Basin</td>
<td>106.7</td>
<td>−512</td>
<td>Duch, River Fm sst &amp; sht</td>
<td>151</td>
<td>1.42</td>
<td>n/a</td>
<td>60%</td>
<td>alluvial</td>
<td>118</td>
<td>4511</td>
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<td>22</td>
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<tr>
<td>lower Uinta Basin</td>
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<td>Green River Fm Ist &amp; sht</td>
<td>134</td>
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<td>122</td>
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<td>18</td>
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<td>upper Colton sst</td>
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<td>n/a</td>
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<td>bedrock</td>
<td>94</td>
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<td>middle Colton sst</td>
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<td>1.42</td>
<td>60%</td>
<td>alluvial</td>
<td>86</td>
<td>696</td>
<td>0.0020</td>
<td>193</td>
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<td>−375</td>
<td>North Horn Ist</td>
<td>127</td>
<td>4.70</td>
<td>n/a</td>
<td>45%</td>
<td>bedrock</td>
<td>85</td>
<td>686</td>
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<tr>
<td>middle Gray Canyon</td>
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<td>−372</td>
<td>Price River Fm sst &amp; sht</td>
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<td>5.17</td>
<td>5.72</td>
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<td>alluvial</td>
<td>78</td>
<td>393</td>
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<td>Blackhawk &amp; Castlegate sst</td>
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<td>bedrock</td>
<td>82</td>
<td>368</td>
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<td>Gunnison Valley</td>
<td>25.4</td>
<td>−352</td>
<td>Mancos shl</td>
<td>8.58</td>
<td>0.49</td>
<td>50%</td>
<td>alluvial</td>
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<td>1076</td>
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<th>Reach Name</th>
<th>Length (km)</th>
<th>River mile</th>
<th>Bedrock/Formation sampled</th>
<th>Compress. strength (MPa)</th>
<th>Tensile strength (MPa)</th>
<th>Alluvial/bedrock</th>
<th>Channel width (m)</th>
<th>Valley width (m)</th>
<th>Gradient (m/m)</th>
<th>Unit stream power (W/m²)</th>
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a Starting river mile of reach, relative to Lee’s Ferry, keyed to nearest mile in Belnap’s River Guides for common reference.
b Effective tensile strength is calculated for alluvial reaches from the regression through the tensile strength–valley width relation for bedrock reaches (Fig. 7).
c Estimated percentage of stratigraphic thickness observed to be comprised of shale, mudrock or other incompetent units.
d Measured width of the valley 50 m above local channel edge.
e Although other units outcrop, the Dox sandstone underlies the vast majority of this reach length.
Fig. 3. Reach-average valley width (measured at 50 m height above channel) as a function of channel width. Black dots are bedrock reaches and gray diamonds are alluvial reaches. Valley width in alluvial reaches does not scale with channel width, whereas there is a positive correlation in bedrock reaches. Bedrock outliers with anomalously high channel width (not included in regression) are reaches 33 and 35 in Glen Canyon, mostly through the massive, weak Navajo Sandstone.

Fig. 4. Box-and-whisker plot of the means and standard deviations of tensile rock strength data illustrating trends as geologic age and degree of burial decrease from left to right. Limestone and basalt are exceptions to the trend of younger, less indurated rocks being weaker.

4. Results

4.1. Basic trends in rock strength

Assessing the database of rock strength, a first-order trend is that older, more deeply buried and indurated rocks are indeed generally stronger than younger rocks, with the exceptions of limestone and Quaternary basalt (Fig. 4; SM Table 1A). This gives us confidence in the dataset, and we expect welded, igneous basalt flows and massive limestones in this desert landscape to be highly resistant to weathering and erosion. Individual tensile strength measurements range from <1 to 15.2 MPa. The lowest corresponding with Mesozoic eolian sandstones of the Wingate, Navajo and Entrada formations, and the highest with the Paleozoic limestones and Quaternary basalt as well as Proterozoic basement and certain highly cemented sandstone beds (SM Table 1A). Schmidt hammer compressive strength values range from 30 to 191 MPa, with the highest values originating from Proterozoic basement rocks and the lowest from friable outcrops of the Permian Hermit and Cambrian Bright Angel shales (SM Table 1B). Selby RMS scores range from 35 to 90 (out of a possible 100); the highest values likewise correspond with Paleozoic limestones and Proterozoic basement rocks, as well as some highly cemented sandstone beds within otherwise shaley units. Some of the massive Mesozoic sandstones have relatively high Selby RMS values because this ranking incorporates their low fracture spacing, whereas the lowest Selby RMS ratings are for the highly fractured and thinly bedded Cambrian mudstones and shales (SM Table 1C).

4.2. Relation of rock strength to fluvial metrics

A first-order research question is whether high bedrock strength corresponds to steep, narrow, high-energy reaches along this river system. High unit stream power and steepness-index values distinguish four knickzones along the system (Figs. 1 and 2), and these values are especially high in Cataract and Grand Canyons where the river encounters Paleozoic carbonates (very resistant to weathering in this arid environment) and Proterozoic crystalline basement. The upstream knickzones have more moderate gradient, occurring where the Green River traverses the Uinta Mountains and in Desolation Canyon, where a relation to rock strength is less clear. Cataract Canyon is an unusual case because it is floored by weak Paradox Formation salt, which creates the regional-scale, translational slide or lateral-spread mass movement of the Needles fault zone (Huntoon, 1988). This salt bedrock is rising diapirically and causing the advection of rock into the canyon (Huntoon, 1988), likely compounding the anomalously steep reach. Because we cannot sample and capture the mechanical strength of dissolving salt, this steepest reach is excluded from our data analysis below. The two reaches upstream of Cataract Canyon are likewise unusual because the river here is interpreted as impounded above the cataracts (Webb et al., 2004), resulting in anomalously low gradient and a disconnect between resistant bedrock canyon walls and fluvial form. These reaches are represented in Fig. 3 by lighter shading and are excluded in the following analysis of bedrock reaches.

Overall, our data confirm that rocks with higher measured tensile strength correspond visually (Fig. 2) and statistically to steeper and narrower reaches with higher unit stream power (Fig. 5, Table 2). The river in Grand Canyon, with a high overall average unit stream power (674 W/m²), runs mostly through hard Proterozoic and Paleozoic rocks with mean and peak tensile strengths of 7.90 and 11.15 MPa, respectively. Though Cataract Canyon is underlain by salt, it has high overall average unit stream power (382 W/m²) and walls of strong Paleozoic rock with a mean tensile strength of 11.44 MPa. Near the upstream end of the transect, the bedrock reaches of the eastern Uinta Mountain knickzone pair similarly hard rock with somewhat lower average unit stream power (293 W/m²). Finally, although Desolation Canyon has both bedrock and alluvial reaches and is the only knickzone that occurs in Mesozoic and Cenozoic rocks, it too matches moderately high stream power to moderately strong rocks (4.56 MPa mean tensile strength). Importantly, no knickzones are present in the least resistant Cenozoic and Mesozoic rocks of the Colorado Plateau.

Statistically, focusing on bedrock reaches, Table 2 indicates the strongest Spearman Rank correlations are between tensile strength and unit stream power, tensile strength and channel width, and valley width and unit stream power. Tensile strength has a stronger

| Table 2 | Spearman rank correlation between bedrock strength and geomorphic metrics. |
|---------------------------------|------------------|--------------------|------------------|------------------|
| Valley width                    | Unit stream power | Tensile strength   | Compressive strength |
| Channel width                   | 0.46             | −0.55*             | −0.69            | −0.64            |
| Compressive strength            | −0.17            | 0.47               | 0.61             |                  |
| Tensile strength                | −0.50            | 0.66               |                  |                  |
| Unit stream power               | −0.73            |                    |                  |                  |

* Unit stream power and channel width are self-correlating values.
correlation than compressive strength to both channel and valley width and also unit stream power, supporting the focus on tensile rock strength as a more relevant measure of erodibility, considering the dominant processes of bedrock stream erosion should be plucking and abrasion in tension. Tensile rock strength has a strong power-law correlation to unit stream power (Fig. 5A) and a linear relation to gradient (Fig. 5B), almost crossing the Y-axis at the origin. These correlations suggest that bedrock strength is the primary control on fluvial and canyon form along this trunk drainage.

4.3. Reach-scale exceptions

At a finer, reach-by-reach scale, correspondence between rock strength and fluvial metrics becomes rich with exceptions, partly due to sampling challenges. For example, reach 13 with low gradient and stream power in the Uinta Basin has an apparent high tensile strength (Fig. 2; Table 1). This is an artifact of sampling bias, where only the more resistant sandstone beds of the shaley Green River Formation are competent enough to be tested. Similarly, hard sandstone beds of the Chinle and Morrison formations misrepresent the overall weak bedrock of mudstone-dominated reaches 8, 10, 19, and 24, as noted in Fig. 2. In a different situation, reach 37 in Marble Canyon has incongruously high stream power and gradient given its moderate tensile strength. This reach is known for the concentration of rapids formed where the river is constricted by tributary debris fans, illustrating how the modern river’s gradient in much of Grand Canyon is set by coarse bed material delivered from adjacent hillslopes by debris flows (Hanks and Webb, 2006). In contrast, lower-than-expected stream power associated with highly resistant rocks is observed downstream in the Lava–Whitmore and Granite Park reaches of western Grand Canyon (reaches 46 and 47). Here the very strong Unkarlet basalt flows occur, but so do weaker pyroclastic materials that cannot be measured, and the valley is notably wider in part due to its alignment along the Hurricane and related faults.

Another example of defying expectations is the massively bedded, eolian sandstones common in the central Colorado Plateau, such as the Navajo sandstone. These rocks have low tensile strength and correspond to wide reaches of low stream power, but they also stand out as having low fracture density (Fig. 2). Theoretically, low fracture density should contribute to low erodibility (especially by plucking) and narrow, steep canyons. However our results show a weak trend of wider fracture spacing corresponding with lower stream power and wider valleys (Fig. 6). Indeed, in the Colorado Plateau, more densely fractured rock is found in the Grand Canyon and Uinta knickzones, and fracture spacing and unit stream power have an inverse correspondence, especially evident through Glen and Grand Canyons (Fig. 2D).

Fracture spacing in our dataset was measured and recorded in the field using the Selby RMS classification into five bins of average width for a given outcrop. Yet, our reach-average values of fracture spacing fell into only the three central Selby RMS bins: 0.05–0.3 m (mean = 0.175, n = 2), 0.3–1.0 m (mean = 0.65, n = 28) and 1.0–3.0 m (n = 1, value = 1.3). Within the 0.3–1.0 m bin, our reach-average data fell into three distinct populations as illustrated in Figs. 2D and 6, with fracture-spacing near the minimum end (spacing = 0.3 m), values central to the bin (0.65 m), and values at the high end (1.0 m). All the sample localities were selected to avoid being near fault zones that could impact the characterization with additional local fractures. Thus, this trend in fracture patterns is representative of rock type and not regional structures, and in this landscape bed rock thickness of sedimentary rocks sets much of the fracturing characteristics. We interpret that the ex-
The rank correlation of tensile strength to valley width in bedrock reaches indicates a moderate correlation (Table 2), whereas strength has more robust correlations with stream power and channel width. Yet, when it comes to alluvial reaches, only valley width varies enough to be useful for projecting an “effective” tensile strength (Fig. 3). Also, it has been shown that there is a process linkage between rock strength and valley width through stream-power driven erosion (e.g. Montgomery, 2004; Limaye and Lamb, 2014), and indeed the correlation of valley width to unit stream power in our dataset is strong (Table 2). Through this process logic we recognize an opportunity to utilize valley width to estimate an “effective” tensile strength (in MPa) of rocks that cannot be sampled representatively or be directly measured (units with >50% shale or other immeasurable rock types). These alluvial reaches and rock units, such as the Green River and Chinle formations mentioned above, are currently represented by the anomalously high strength values of only the fraction of outcrop that was resistant enough to withstand sampling and strength testing.

In bedrock reaches, tensile-strength varies over an order of magnitude, and it exhibits a rough correlation with valley width (Fig. 7), with resistant rock units underlying narrower canyons and weaker formations associated with wider valleys. Reaches 38, 42, 44, and 45, all within Grand Canyon, form a distinct group of outliers (circled in Fig. 7), which have the narrowest valley-bottoms in the study but tensile strengths that are not significantly higher than somewhat wider reaches. The Redwall and Muav limestones of middle Marble Canyon and the Muav Gorge (reaches 38 and 45) form nearly vertical walls and entrenched bedrock meanders, and they are probably even more difficult to erode in this arid setting than their tensile strength values suggest. Likewise, the measured tensile strengths for the Zoroaster granite and Vishnu schist basement rocks (reaches 42 and 44), albeit high, fail to capture how narrow those inner gorges are. We suggest the strength of these variable and foliated rocks is underestimated with laboratory testing methods. Regardless, projecting for alluvial reaches, our “effective” tensile-strength values increase the total range to three orders of magnitude for the rock units present in the Colorado Plateau (Fig. 7 and Table 1). This is consistent with the range of tensile strength values presented by Sklar and Dietrich (2001), which includes artificial mixtures made to represent weaker rocks.

5. Discussion

5.1. Correspondence of rock strength and fluvial morphology

Our first-order research question has been whether strong bedrock does or does not correspond to steep, narrow, high-energy reaches along this river system. They do. We document trends of older rocks being more resistant than younger rocks, as well as correlations between tensile strength and river gradient, channel width and unit stream power of bedrock reaches. Interesting exceptions exist in our dataset, which we have interpreted above as the product of specific lithologic, weathering and geologic variations from reach to reach. Thus, rock strength should not be ignored in landscape evolution and we confirm that simple correlations between topographic and long-profile metrics and incision or uplift rate are unlikely to exist outside of steady-state landscapes. Furthermore, finding and interpreting topographic signals of differential uplift, transient incision, or alluvial bed-cover effects becomes that much more complicated.

We offer our dataset as a potential aid for parameterizing landscape evolution models. The work of Sklar and Dietrich (2001) indicates that K scales to the inverse square of tensile rock strength, which implies that K should span five orders of magnitude for our (three order-of-magnitude) range of measured and effective tensile strengths. For comparison, the modeling studies of DeLong et al. (2007) and Pelletier (2010) set a range of values of K over less than three orders of magnitude for particular study areas, which do not span a suite of lithologies as variable as this study. Stock and Montgomery (1999), who worked with incision of a broader set of granitic, metamorphic, volcaniclastic and mudstone rocks in Australia and Hawaii, found values of K ranging from $10^{-2}$ m$^{0.2}$/yr to $10^{-7}$ m$^{0.2}$/yr, a range as large as we suggest.

Bedrock properties are certainly recognized as a primary control of channel form for bedrock streams, as is sediment load or bed cover. The modeling work of Yanites and Tucker (2010) suggests that channel width is a key adjustment that depends upon sediment load. Specifically, as sediment and bed cover is increased, channel width increases significantly and gradient steepens slightly. This model result raises the question: to what degree can the variations in the sediment-rich Colorado River’s long profile be attributed to sediment-cover effects rather than bedrock strength? In this study, especially in Glen and Grand Canyons, steeper reaches are narrower, not wider, which is opposite the trend suggested by Yanites and Tucker (2010) for the bed-cover effect. Inasmuch as our dataset couples steepness with narrowness, a trademark of streams in communication with bedrock, there is evidence that the Colorado River’s large-scale long profile, especially in canyon knickzones, is set more by bedrock than sediment-cover effects.

A related issue was raised by Hanks and Webb (2006), who suggested that the river profile, specifically through Grand Canyon, is controlled by thick accumulations of Holocene debris flows and fans forming two multi-reach convexities along the channel profile. Individual rapids occur where the channel is constricted with
debris from local tributaries. In this study we find, however, that most reach-scale differences in gradient, channel width and stream power in Grand Canyon correspond to changes in bedrock strength, including the low gradient upper reaches versus the steeper downstream reaches of both of Hanks and Webb’s (2006) convexities. Rather than direct channel contact with bedrock everywhere, a resolution may lie in an indirect effect, whereby hillslopes and tributary canyons supply bedload of a size and hardness representing the local bedrock.

5.2. Implications for Colorado Plateau landscape evolution

After the integration of the Colorado Plateau across the southwestern edge of the Colorado Plateau dropped baselevel ~6 million years ago, upstream incision has been non-uniform, as the highly uneven long profile of the Colorado River across the region exemplifies (Fig. 2). Understanding the processes responsible for this irregular river profile and the uneven distribution of erosion will shed light on the active debates about regional uplift and erosion. Pederson and Tessler (2012) suggested that bedrock resistance is a primary controlling factor on not only the Green–Colorado long profile, but also the river’s width and thus unit stream power (Figs. 2 and 7). We confirm this with rock-strength data, and reiterate that much of this river’s highly variable long profile is therefore an expression of dynamic equilibrium: where gradient and width are adjusted to match substrate resistance.

Alternative hypotheses to explain the variable knickzones of the Colorado River include transient baselevel signals and differential uplift (Cook et al., 2009; Crow et al., 2014). If knickzones represent transient features, that implies spatial differences in incision rates and a degree of disequilibrium where bedrock strength would not necessarily match river steepness or other metrics. Cook et al. (2009), for example, suggest that the higher incision rates upstream of Lee’s Ferry versus downstream in Grand Canyon supports such a transient state, and that transient knickpoints may be found along smaller tributaries. Although transient features may exist in the tributaries of the greater drainage, our data suggest that an equilibrium with bedrock resistance plays a dominant role in the trunk–river profile, with the caveat that the Desolation knickzone has a weaker correspondence to rock strength (Pederson and Tessler, 2012). Furthermore, isotopic feedback accounts for much of the higher incision rates in the central Colorado Plateau (Pederson et al., 2013). The differential-uplift alternate hypothesis focuses on the Grand Canyon knickzone and an explanation linked to flow in the upper mantle. Crow et al. (2014) suggest this may have driven 0.5–1.5 km of differential surface uplift of the southwestern edge of the Colorado Plateau since 10 Ma forming the Grand Canyon knickzone. However, our correlations between rock strength and steepness are able to account for the gradient of the river more simply and remove it as evidence for differential mantle uplift.

6. Conclusions

Our empirical data on rock strength include reassuring findings that tensile and compressive strength correspond and that rock strength varies systematically and logically with age (burial) and rock type. Relative to fluvial topography, reach-scale, averaged tensile strength correlates positively with unit stream power and inversely to channel width with power-law relations and correlates with gradient in a linear relation. Stream power also inversely correlates with valley width taken at 50 m above local channel elevation, representing the long-term ability of the channel to planate the valley bottom. Reach-scale exceptions to these relations can be linked to non-representative sampling of shaley units and other immeasurable materials or to particular weathering and mechanical variations, such as the massive but weak eolian sandstones of the central plateau. These correlations of fluvial metrics to variable bedrock, and through tensile strength, must be considered in efforts to use such metrics to interpret tectonics in settings around the world. This dataset may be useful for constraining models of landscape evolution, and we suggest that our estimates of effective tensile strength, as scaled to valley width and ranging over five orders-of-magnitude, may be a solution to the problem of how to incorporate rock types too weak or variable to capture by sampling.

Our results reaffirm that bedrock strength is a first-order control on large-scale fluvial geomorphology, especially in the non-steady-state, arid, and geologically variable terrain of the Colorado Plateau. Recognizing this regional control renders secondary some more complicated alternate explanations involving differential mantle uplift, transient incision, and bed-cover effects. In this iconic landscape of steep canyons and broad plateaus, John Wesley Powell had it right from the start regarding the controls of bedrock on geomorphology.

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Appendix A. Supplementary material

Supplementary material related to this article can be found online at http://dx.doi.org/10.1016/j.epsl.2015.07.042.

References


